

Observation of Water Vapor Greenhouse Absorption Over the Gulf of Mexico Using Aircraft and Satellite Data

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ABSTRACT

Through its interaction with radiation, water vapor provides an important link between the ocean and atmosphere. One way this occurs is through the greenhouse effect, and here we report on observations of water vapor greenhouse absorption in the Gulf of Mexico during the Cirrus Regional Study of Tropical Anvils and Cirrus Layers – Florida Area Cirrus Experiment during July 2002. The quantity dG_a/dT_s is the change in the amount of upwelling infrared flux absorbed by water vapor as the sea surface temperature increases, and therefore parameterizes the strength of the evaporative feedback between the ocean and atmosphere. Using hemispherical (IRBR) and narrow field of view (NFOV) radiometers aboard a NASA ER-2 aircraft, we have measured dG_a/dT_s during flights on July 9 and July 26 marked by large scale convective and quiescent conditions, respectively. Using the NFOV over the wavelength range $4 - 40 \mu\text{m}$, we measured $dG_a/dT_s = 13.4 \pm 1.0 \text{ W m}^{-2} \text{ K}^{-1}$ on July 9, while on July 26 we measured $dG_a/dT_s = 9.7 \pm 0.3 \text{ W m}^{-2} \text{ K}^{-1}$. NFOV measurement of dG_a/dT_s in the $8 - 12 \mu\text{m}$ wavelength range yielded values of $\sim 2.5 \text{ W m}^{-2} \text{ K}^{-1}$ for both days, indicating that most of the change in greenhouse absorption with increasing ocean temperature occurs in the rotational and vibrational spectral regions of water vapor. IRBR measurements yielded higher values of dG_a/dT_s on both days, but were likely affected by cold clouds in the hemispherical radiometer field of view. These results support the link between greenhouse efficiency, mid to upper tropospheric water vapor concentration, and convection.

1. Introduction

Water vapor is the most important atmospheric greenhouse gas due to its ubiquity and its molecular absorption bands in the infrared. In the absence of dynamic effects, the heat-absorbing effect of water vapor could result in runaway heating of the tropical ocean due to radiative feedbacks. This can be seen as follows. As the ocean is heated by solar insolation, a significant fraction ($\sim 40\%$; Raval and Ramanathan, 1989) of the outgoing longwave radiation is absorbed by water vapor in a layer < 4 km above ocean surface. This optically thick layer then re-radiates back onto the ocean surface, resulting in ocean heating (VonderHaar, 1986; Lubin, 1994). This heating would then result in an exponential increase in water vapor density due to the sensitive dependence of the saturation vapor pressure on temperature (Raval and Ramanathan, 1989; Stephens, 1990), resulting in further heating. In some areas of the tropics, runaway heating of the ocean surface would then result if this process proceeded unchecked (the “super greenhouse effect”; VonderHaar, 1986). It is estimated that over 50% of the tropical sea surface during all seasons is subjected to strong ocean heating by super greenhouse conditions (Valero et al., 1997a).

This thermal runaway does not happen because the ocean and atmosphere are a coupled system (e.g. Waliser and Graham, 1993), and dynamical effects such as cloud feedback (Ramanathan and Collins, 1991; Miller, 1997; Collins et al., 2000), evaporative cooling (Newell 1979; Fu et al., 1992; Hartmann and Michelsen, 1993; Tsintikidis and Zhang, 1998), and circulation in the ocean (Sun and Liu, 1996; Clement and Seager, 1999) and atmosphere (Graham and Barnett, 1987; Wallace, 1992; Fu, Del Genio, and Rossow, 1994; Pierrehumbert, 1995) all conspire to smooth out gradients in sea surface temperature on long timescales. A crucial link in this process is water vapor, because water vapor both cools the ocean – through evaporation and shortwave cloud forcing – and heats the ocean through both cloud and water vapor longwave emission. The tropical water vapor feedback

mechanism affects global climate directly due to the observed positive correlation between the mean temperature of the tropical oceans and global tropospheric temperatures (Newell and Weare, 1976; Angell 1990). This feedback is also important because it can amplify global warming due to increased anthropogenic carbon dioxide emission by a factor of two over global warming projections with no water vapor feedback (IPCC Report, 2001).

The variation in the water vapor greenhouse absorption with sea surface temperature yields the total efficiency at which water vapor heats the ocean. We therefore use the term “greenhouse efficiency” to refer to the observed rate of change in greenhouse absorption with changing sea surface temperature. The atmospheric greenhouse absorption G_a is commonly defined by (Ramanathan and Collins, 1991; Hallberg and Inamdar, 1993):

$$G_a = F_B(T_s) - F_{\uparrow}, \quad (1)$$

where F_{\uparrow} is the observed upwelling infrared flux above the water vapor layer, T_s is the sea surface temperature in Kelvin, and the blackbody flux from the ocean surface is $F_B = \sigma T_s^4 \eta$. Here $\sigma \approx 5.67 \times 10^{-8} \text{ W m}^{-2} \text{ K}^{-4}$, and the dimensionless Planck integral is given by

$$\eta = \frac{15}{\pi^4} \int_{x_1}^{x_2} \frac{x^3 dx}{e^x - 1}, \quad (2)$$

where $x = 10^4 hc/(\lambda_i k T_s)$, λ is the wavelength in microns and $h \approx 6.63 \times 10^{-27} \text{ erg s}$ and $c \approx 3.00 \times 10^{10} \text{ cm s}^{-1}$ are Planck’s constant and the speed of light, respectively. Assuming $T_s = 300\text{K}$, $\eta \approx 0.94$ for the broadband infrared spectral region with $\lambda_2 = 4.0 \text{ }\mu\text{m}$ and $\lambda_1 = 40.0 \text{ }\mu\text{m}$, and $\eta \approx 0.26$ for the atmospheric window region of the spectrum, with $\lambda_2 = 8.0 \text{ }\mu\text{m}$ and $\lambda_1 = 12.0 \text{ }\mu\text{m}$.

The quantity dG_a/dT_s gives both the change in upwelling flux absorbed by the water vapor layer *and* the change in downwelling flux radiated onto the ocean with increasing sea surface temperature, and therefore parameterizes the strength of the ocean-atmosphere coupling and the water vapor feedback mechanism. Using satellite data, Raval and

Ramanathan (1989) and Ramanathan and Collins (1991) showed that $dG_a/dT_s \sim 5 \text{ W m}^{-2} \text{ K}^{-1}$ for $T_s < 298 \text{ K}$ but increased to $dG_a/dT_s \sim 10 \text{ W m}^{-2} \text{ K}^{-1}$ for $T_s > 298 \text{ K}$ over the central Pacific warm pool during April 1987. Also using satellite data, Hallberg and Inamdar (1993) showed that the efficiency of greenhouse absorption varied significantly across the Globe and with the season, with the highest values of dG_a/dT_s associated with convective regions. Lubin (1994) observed downwelling fluxes from a research vessel in the central Pacific, and inferred values of dG_a/dT_s consistent with the previously mentioned satellite measurements. Finally, additional observations of dG_a/dT_s over the central Pacific warm pool region from a NASA ER-2 aircraft (Valero et al., 1997a) found $dG_a/dT_s = 14 - 15 \text{ W m}^{-2} \text{ K}^{-1}$, with the higher values again associated with convective regions and sea surface temperatures greater than 301 K.

Here we present aircraft and satellite observations of the water vapor greenhouse efficiency over the Gulf of Mexico during July 2002. The data consist of clear sky observations from a suite of downward-looking radiometers aboard a NASA ER-2 aircraft and the Terra satellite. This paper is organized as follows. In section 2 we discuss our radiometric observations of dG_a/dT_s in the Gulf of Mexico, we present our results in section 3, and a discussion in section 4. Finally, we summarize our results in Section 5.

2. Observations

These observations were conducted during the NASA Cirrus Regional Study of Tropical Anvils and Cirrus Layers – Florida Area Cirrus Experiment (CRYSTAL-FACE) in July 2002. During this mission, a narrow field of view (NFOV) and a broadband (IRBR) infrared radiometer were flown aboard a NASA ER-2 aircraft as part of the Radiation Measurement System (RAMS; Valero et al. 1982). The NFOV consists of two independent detectors sensitive over the wavelength ranges $4 - 40 \mu\text{m}$ and $8 - 12 \mu\text{m}$ and collimated to an 8°

downward-looking field of view (FOV). During flight, the NFOV detectors are calibrated in real time by chopping between the downward-looking scene and a cryogenically-controlled blackbody fifteen times a second, and an absolute radiance calibration is obtained using a liquid blackbody source in the lab. The IRBR is a $4 - 40 \mu\text{m}$ hemispherical FOV radiometer. As with the NFOV, the IRBR is calibrated in the lab using a liquid blackbody source. Details on the calibration of the RAMS infrared radiometers are given elsewhere (Valero et al., 1997b). The NFOV is mounted in an external pod located on the lower port side of the ER-2, and the IRBR is located in the System-20 (S20) pod on the ER-2 starboard wing. The tip of the S20 pod rotates throughout the flight, switching the FOV of the IRBR between the up and down position once every 30 seconds¹.

During CRYSTAL-FACE, the ER-2 containing the RAMS payload flew out of Key West Naval Air Station near Key West, Florida. While most of the flights were devoted to observations of convective clouds near Florida ground stations, the ER-2 flights on July 9 and 26, 2002, were “Southern Survey” flights in which the aircraft flew south over the Gulf of Mexico and Caribbean Sea to latitude 14° before returning to Key West. During these flights, stretches of clear sky data were observed during the return legs when the ER-2 was approaching Key West over the Gulf of Mexico. In the area of Key West, July 9 2002 was marked by large scale convection, while July 26 was characterized by a relatively stable, convection-free boundary layer. The ER-2 altitude during the time of the two observing intervals were 21.4 and 20.2 km for July 9 and 26, respectively, and portions of the flight tracks for the two days are shown in Figures 1 and 2. For comparison, in Figure 1 we have overplotted contours of the 10 m wind speed as measured by the SeaWinds scatterometer aboard the QuikScat satellite (Dunbar, Hsiao, and Lambrigtsen, 1991). Satellite-derived maps of sea surface temperature at a temporal and spatial resolution of 48 hr and 14 km,

¹The actual data accumulation time per position is $2/3$ of this, or 20 seconds/cycle.

respectively, were obtained from the NOAA Satellite Active Archive for July 26, 8, and 10, 2002. The sea surface temperature maps from July 8 and 10 were averaged to obtain a map for July 9, and the use of sea surface temperature data from July 24 or 28 in our analysis (instead of the July 26 data) would not have changed our final results significantly for that day. The absolute accuracy of the satellite-derived sea surface temperatures is $0.5 - 1.0$ K, although we are mainly concerned with the relative precision of the data, which is better than 0.1 K (Valero et al., 1997a).

Comparisons of satellite and in-situ measurements are important for validation of global scale studies that use satellite data exclusively. Therefore we use upwelling longwave fluxes at the top of the atmosphere (TOA) as measured by the Clouds and the Earth’s Radiant Energy System (CERES) instrument aboard the Terra satellite. Terra is a polar-orbiting satellite with an orbital inclination of 98.2° . The CERES Terra data we use here consist of upwelling longwave fluxes and geolocated latitude and longitude data from the CERES Terra FM1 Edition1 ES8 archive. CERES data is geolocated independently of the cloud-viewing scene. An angular distribution model (ADM) is used to convert the radiances measured by CERES into upwelling $5.0 - 100.0 \mu\text{m}$ longwave fluxes, with a total uncertainty of 12% primarily due to the ADM (Wielicki et al., 1996). The CERES Terra data were collected 3 hours before the RAMS ER-2 data on both July 9 and July 26, and matched to the ER-2 flight track using bilinear interpolation. Greyscale images of the non cloud-screened CERES data are overplotted shown Figures 1 and 2. Only data marked as clear by the CERES cloud detection algorithm (Wielicki et al., 1996) were used in this analysis. Unfortunately this excluded most of the CERES data from July 9, as the satellite image for that day (Figure 1) shows a thin cloud over the latter half of clear-sky ER-2 track. For the July 26 observations, however, there were enough CERES data after cloud screening to provide a useful comparison with the ER-2 data.

3. Results

3.1. Observed Radiances and Model Comparison

The upwelling $4 - 40 \mu\text{m}$ and $8 - 12 \mu\text{m}$ radiances from the NFOV as a function of time are shown for the July 9 flight and the July 26 flight in Figures 3-4. Also shown in the same figures are the $4 - 40 \mu\text{m}$ upwelling fluxes from the IRBR. All the data has been averaged over the 20 s dwell time of the S20 pod. Clear sky regions are indicated in Figures 3 and 4; these were determined using preliminary images from the Cloud Physics Lidar (CPL: McGill et al., 2002) and Modis Airborne Simulator (MAS: King et al., 1996) instruments aboard the ER-2. Inspection of the preliminary $1.06 \mu\text{m}$ CPL and $11.01 \mu\text{m}$ MAS images for July 9 showed no evidence of significant cirrus or boundary layer clouds directly under the aircraft during the designated clear intervals, but for the July 26 data we additionally exclude NFOV data from the interval 20.41 hr to 20.47 hr GMT, during which the ER-2 passed directly over a cirrus layer. A small dip in the $4 - 40 \mu\text{m}$ NFOV upwelling radiance is apparent during this interval in Figure 4, and an even longer dip is visible in the IRBR data taken during this time. This is due to clouds adjacent to the NFOV field of view but seen by the hemispherical IRBR radiometer, and we therefore exclude additional IRBR data from 20.28 hr to 20.41 hr GMT on July 26 from our analysis.

We model the observed NFOV radiances for two representative times during each flight using SBDART V2.0 (Ricchiazzi, Yang, and Gautier, 1998), a plane-parallel radiative transfer routine based on the discrete ordinates method (Stamnes et al., 1988). Input to the model were the interpolated sea surface temperature for the given time, and the atmospheric pressure, temperature, and water vapor density from ER-2 dropsonde measurements (Hock and Franklin, 1999) for altitudes less than 17 km, and a standard tropical atmosphere profile for altitudes greater than this. We use data from two sondes dropped from the ER-2

during the two flight legs analyzed in this paper, yielding total water vapor column densities of 5.76 g cm^{-2} and 4.98 g cm^{-2} for the July 9 and 26 flights, respectively. For comparison with data radiances from the 8° NFOV field of view, model radiances were calculated for 10 zenith and 19 azimuth angles uniformly spaced from $0^\circ - 4^\circ$ about the vertical upwelling direction and averaged using numerical integration. The comparison between the data and model calculations for the two test cases are shown in Table 1, where the model upwelling radiances have been interpolated to the altitude of the ER-2. The model and observed radiances are within 3% for all the test cases. More frequent measurements of the vertical distribution, and integrated amount, of water vapor would be much more desirable than the two points modeled here, but unfortunately such data were not available on the ER-2.

Using the model results in Table 1, we can derive a conversion factor between the NFOV radiances and the upwelling flux, defined by (Collins et al., 2000)

$$F_{\uparrow} = \pi l I_{\uparrow}, \quad (3)$$

where I_{\uparrow} is the upwelling radiance, and l is the conversion factor. For an isotropic radiance distribution $l = 1$, but limb-darkening effects produce $l < 1$ for realistic radiances. Using the radiative transfer calculation in Table 1, we derive $l_{4-40} = 0.935$ and $l_{8-12} = 0.930$ for the $4 - 40 \text{ } \mu\text{m}$ and $8 - 12 \text{ } \mu\text{m}$ data on July 9, and $l_{4-40} = 0.941$ and $l_{8-12} = 0.945$ for the July 26 data.

3.2. Calculation of the Greenhouse Efficiency

For the clear sky intervals in Figures 3-4, the greenhouse absorptions were calculated for the NFOV and IRBR using equation (1). The radiometer data were binned on 20 second intervals, and the corresponding sea surface temperatures were derived via bilinear interpolation. The results for $4 - 40 \text{ } \mu\text{m}$ are shown in Figures 5 and 6 for the July 9

and 26 data, respectively. The measured values of G_a on July 9 are systematically higher than the July 26 values, due to the presence of more water vapor higher up in the atmosphere on July 9 (as shown in Table 1 and depicted in Figure 8 discussed below). A linear regression of G_a vs T_s produces the slopes given in these figures and Table 2, with the error on the slope calculated from the root mean square residuals of the linear regression. The linear fits to the $4 - 40 \mu\text{m}$ data are highly significant, with correlation coefficients greater than 0.9. For both days the fluxes from the hemispherical IRBR yield values of dG_a/dT_s that are $\sim 30 - 50\%$ steeper than the corresponding values derived from the NFOV radiances, which is likely due to the presence of cold clouds in the hemispherical field of view of the IRBR (see Section 4). For the $5 - 100 \mu\text{m}$ CERES Terra data, a greenhouse efficiency of $14.9 \pm 0.6 \text{ W m}^{-2} \text{ K}^{-1}$ was measured along the ER-2 flight tracks three hours before the RAMS data was taken on July 26. The offset between the CERES and ER-2 points is due at least in part to the larger $5 - 100$ micron bandpass of the CERES instrument, absorption by ozone at altitudes greater than 20 km, and field of view effects discussed below.

Figures 7 and 8 show the same plots for the $8 - 12 \mu\text{m}$ channel of the NFOV radiometer for the two flights. The fitted values for the slope dG_a/dT_s are $2.4 \pm 0.7 \text{ W m}^{-2} \text{ K}^{-1}$ and $2.6 \pm 0.2 \text{ W m}^{-2} \text{ K}^{-1}$ for the July 9 and July 26 data, respectively. Compared to the $4 - 40 \mu\text{m}$ data, the correlation of G_a with T_s is not as significant for the $8 - 12 \mu\text{m}$ data, with values of the correlation coefficient given in Table 2. We detect no significant change in the $8 - 12 \mu\text{m}$ greenhouse efficiency between the two flights, although a 50% increase in $8 - 12 \mu\text{m}$ dG_a/dT_s between July 9 and July 26 is consistent with the data at the 2σ level. The mean value of the greenhouse efficiency for $8 - 12 \mu\text{m}$ is $2.6 \pm 0.2 \text{ W m}^{-2} \text{ K}^{-1}$, corresponding to $\sim 19\%$ and $\sim 27\%$ of the $4 - 40 \mu\text{m}$ NFOV greenhouse efficiency during the July 9 and 26 flights, respectively.

4. Discussion

The broadband greenhouse efficiency we obtain on July 9 is consistent with similar aircraft measurements of dG_a/dT_s over the central Pacific warm pool region in 1993 (Valero et al., 1997a). Both of these results are associated with large scale convection, and the lower dG_a/dT_s value we obtain on July 26 occurred during largely convection-free conditions despite the higher mean sea surface temperature during this flight. This supports the observational (Soden and Fu, 1995) and theoretical (Allan et al., 1999) assertion that atmospheric water vapor concentration, upper and middle tropospheric humidity, and convection, but not necessarily high sea surface temperatures for $T_s > 300$ K (Graham and Barnett, 1987), are associated with high values of the water vapor greenhouse efficiency. Our measurements of much smaller values of dG_a/dT_s in the water vapor continuum region also support this hypothesis, as they indicate that most of the greenhouse absorption must occur in the rotational and vibrational absorption regions of the infrared spectrum. Since these absorption bands are mostly saturated in the very humid air of the tropical lower troposphere (Allan et al., 1999), the change in greenhouse absorption must take place in the middle or upper troposphere, where the water vapor density is relatively low. This is also supported by water vapor density profiles observed by the ER-2 dropsonde measurements on July 9 and 26; a plot of the ratio of H₂O density as a function of height (Figure 9) indicates that the largest change in water vapor density between the two days occurred in the middle to upper troposphere, at altitudes of $z \sim 5 - 10$ km (pressure $\sim 200 - 500$ mb). This supports the link between middle and upper tropospheric moisture and super greenhouse absorption.

Our observations also reveal that higher values of the greenhouse efficiency are obtained by hemispherical field of view radiometers than by narrow field of view radiometers. This effect was also noticeable during the Central Equatorial Pacific Experiment (Valero et al.,

1997a). It seems likely that the discrepancy between the hemispherical FOV measurements and the narrow FOV measurements is due to their differing fields of view. On both of the CRYSTAL-FACE days used in this analysis, CPL images (McGill et al., 2002) indicate that the ER-2 overflow cirrus layers just after the clear intervals indicated in Figures 3-4. This did not affect the NFOV because of its narrow field of view, but the clouds would have grown progressively larger in the hemispherical field of view IRBR toward the end of the flight, when the ER-2 overflowed the warmest sea surface temperatures. Since the high clouds are much colder than the sea, this would have suppressed F_{\uparrow} and therefore increased G_a towards the end of both flights, yielding artificially higher values of dG_a/dT_s as measured by the IRBR. For this reason the greenhouse efficiencies measured by the NFOV instrument are probably closer to the true cloud-free values for these data sets. Clearly measurements of the upwelling flux through a very small field of view are necessary to obtain accurate measurements of the water vapor greenhouse efficiency.

On July 26, the greenhouse efficiency obtained by CERES Terra was similar to but slightly higher than the IRBR and NFOV measurements obtained on that day. The slight discrepancy could be due to the different spectral bandpasses of CERES and RAMS IRBR and NFOV channel 1, changing conditions along the ER-2 flight track in the 3 hours between the CERES Terra and RAMS observations, or possibly to a field of view effect similar to the one discussed above. In the latter case, the much larger CERES pixel size (20 km) compared to that of the NFOV (~ 4 km) could be the cause of the discrepancy. Although cloud screening was used for the CERES data points, the possible presence of thin cirrus towards the end of the ER-2 flight track (Figure 2) would affect the CERES data more than the NFOV due to the larger angular sensitivity of the CERES pixel. In addition, the the ER-2 flight track was near the edge of the CERES scan for the July 26 observation, resulting in a very large zenith viewing angle of $\sim 65^\circ$ from the satellite. The CERES radiance to flux conversion factor is largest for such large viewing angles

(Wielicki et al., 1996), accentuating uncertainties in the angular distribution function and resultant flux errors. For all these reasons, the apparent discrepancy between the RAMS and CERES Terra results on July 26 is not surprising: near contemporaneous satellite and aircraft measurements are needed for adequate intercomparisons between the satellite and in situ platforms. Despite these difficulties, however, the closeness of the ER-2 and satellite measurements of dG_a/dT_s on this day bode well for future studies of the global distribution of greenhouse forcing from space.

5. Summary

We have measured the change in greenhouse absorption as a function of sea surface temperature – the greenhouse efficiency – over the Gulf of Mexico during July 2002 using infrared radiometers aboard both aircraft and satellite. Data from two days were presented in which the amount of atmospheric water vapor varied significantly; one day (July 9) was characterized by large scale convection, while the other day (July 26) was largely non-convective. We find higher broadband greenhouse efficiencies during the convective day while the greenhouse efficiencies in the spectral window region remained essentially unchanged, at one-fourth the broadband value, between the two days. This supports the hypothesis that convection and upper tropospheric moisture are the key determinants in the value of the greenhouse efficiency. Although the aircraft observations of water vapor greenhouse absorption presented here are of short duration and small geographical scope, the consistency between the satellite and aircraft results can be seen as an in situ validation of the satellite data – opening the door for more extensive studies using global satellite observations of tropical water vapor greenhouse absorption and its relation to atmospheric structure, stability, and other factors. Such studies are needed to fully understand the important radiative feedback between the ocean and atmosphere.

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Figure 1: Segment of the flight of the NASA ER-2 during CRYSTAL-FACE on July 9, 2002. The bold line corresponds to the data stretch marked “clear sky” in Figure 3, and similarly with the circled letters along the ER-2 flight path. Overplotted in gray shades are upwelling $5 - 100 \mu\text{m}$ fluxes measured by the CERES instrument aboard the Tera satellite, with the flux in W m^{-2} indicated by the vertical bar on the left. Contours also depict the wind speed (m s^{-1}) as measured on the afternoon of July 9 by the SeaWinds scatterometer aboard the QuikScat satellite. The black areas in the lower left and upper right corners are regions with no CERES data.

Figure 2: Same as Figure 1, but for the ER-2 flight on July 26, 2002. There were no satellite wind observations of this area during this day.

Figure 3: Data from the RAMS radiometer package on the NASA ER-2, from July 9, 2002. The $4 - 40$ and $8 - 12 \mu\text{m}$ upwelling radiances from the NFOV radiometer are shown along with the upwelling $4 - 40 \mu\text{m}$ flux from the IRBR (hemispherical field of view). The clear sky interval, corresponding to the bold line in Figure 1, was determined from Cloud Physics Lidar (McGill et al., 2002) and MODIS Airborne Simulator (King et al., 1996) images taken from the ER-2 during the flight. The circled letters above the time axis correspond to intervals along the ER-2 flight path shown in Figure 1.

Figure 4: Same as Figure 5. but for the July 26, 2002 data.

Figure 5: The $4 - 40 \mu\text{m}$ greenhouse flux from July 9, calculated according to (1), for the NFOV and IRBR radiometers. The NFOV $4 - 40 \mu\text{m}$ radiances have been converted to flux using a conversion factor derived using a radiative transfer model (see text). The indicated values denote the greenhouse efficiency calculated via linear regression of G_a versus sea surface temperature T_s . The 1σ errors on the efficiency are determined from the dispersion of the points about the linear fit.

Figure 6: Same as Figure 5, but for the July 26 data. Also shown are the $5 - 100$ μm fluxes from the CERES instrument aboard the Terra satellite recorded approximately 3 hours before the ER-2 data.

Figure 7: The $8 - 12$ μm greenhouse flux from July 9, calculated according to (1), where the upwelling NFOV radiance has been converted to flux using a model-derived conversion factor as before. The greenhouse efficiency and error were calculated as in Figures 5 and 6.

Figure 9: The ratio of July 9 to July 26 water vapor density as a function of height, as determined from dropsonde measurements from the ER-2. The change in water vapor density is greatest at altitudes of $5 - 10$ km, supporting the link between middle tropospheric moisture and super greenhouse absorption.

Table 1. Comparison of Model and NFOV Upwelling Radiances ¹

| Quantity | Units | July 9 | July 26 |
|----------|----------------------|--------|---------|
| Time | GMT hr | 19.73 | 20.15 |
| Lat. | Deg. | 22.68 | 24.57 |
| Lon. | Deg. | 273.85 | 278.30 |
| z | km | 21.41 | 20.18 |
| T_s | K | 302.0 | 302.0 |
| w | g cm^{-2} | 5.76 | 4.98 |
| NFOV | $4 - 40 \mu\text{m}$ | 90.6 | 93.3 |
| Model | $4 - 40 \mu\text{m}$ | 88.1 | 92.1 |
| NFOV | $8 - 12 \mu\text{m}$ | 33.2 | 35.9 |
| Model | $8 - 12 \mu\text{m}$ | 34.1 | 35.5 |

¹Comparison of observed upwelling radiances and radiative transfer model (SBDART) calculations using indicated inputs and a 39 level atmospheric profile derived from ER-2 dropsonde data (Hock and Franklin, 1999) with integrated water amount w . The SBDART results were interpolated to the ER-2 altitude z .

Table 2. Greenhouse Forcing Efficiency^a

| Inst. | Wavelength (μm) | Day (m/d/y) | T_s Range (K) | dG_a/dT_s ($\text{W m}^{-2} \text{K}^{-1}$) | R |
|-------|---------------------------------|----------------------|--------------------|--|------|
| NFOV | 4.0 – 40.0 | 7/09/02 | 301.7 – 302.2 | 13.4 ± 1.0 | 0.93 |
| IRBR | 4.0 – 40.0 | 7/09/02 | 301.7 – 302.2 | 18.6 ± 1.1 | 0.98 |
| NFOV | 8.0 – 12.0 | 7/09/02 | 301.7 – 302.2 | 2.4 ± 0.7 | 0.54 |
| NFOV | 4.0 – 40.0 | 7/26/02 | 302.0 – 303.2 | 9.7 ± 0.3 | 0.95 |
| IRBR | 4.0 – 40.0 | 7/26/02 | 302.0 – 303.2 | 13.2 ± 0.8 | 0.94 |
| CERES | 5.0 – 100.0 | 7/26/02 ^b | 302.0 – 303.2 | 14.9 ± 0.6 | 0.92 |
| NFOV | 8.0 – 12.0 | 7/26/02 | 302.0 – 303.2 | 2.6 ± 0.2 | 0.86 |

^aSlope dG_a/dT_s from linear regression of greenhouse flux G_a versus sea surface temperature T_s . The correlation coefficient of the fit is R and the slope error is determined from the root-mean-square scatter of the points.

^bCERES data were taken 3 hours before the RAMS data.

















